Millennial-scale climate change and intermediate water circulation in the Bering Sea from 90 ka: A high-resolution record from IODP Site U1340


1. Introduction

[2] Abrupt and millennial-scale climate changes for the last 90 kyr, including Dansgaard-Oeschger (DO) events, the Bølling-Allerød (BA), and the Younger Dryas (YD) [Dansgaard et al., 1993], are thought to result from complex interactions between ice sheets, sea ice, ocean circulation, and atmospheric greenhouse gases [for a review, see Clement and Peterson, 2008]. The global nature of these events became clear when California margin cores provided evidence that North Pacific benthic oxygenation was synchronous with Greenland DO events [Behl and Kennett, 1996; Cannariato and Kennett, 1999]. Studying these oxygenation fluctuations will help to reveal the processes that connect Atlantic and Pacific millennial-scale climate events.

[3] Hendy and Pedersen [2005] found that both export production and intermediate water ventilation influenced changes in benthic oxygenation. Intermediate water at the California margin is affected by the distal extent of two water masses [Hendy and Kennett, 2000]: the oxygen-rich North Pacific Intermediate Water (NPIW) advected from the northwest at 300–500 m [Talley, 1993] and the low-oxygen California Undercurrent advected from the south at 100–500 m [Hendy et al., 2004; Gay and Chereskin, 2009]. However, oxygen variations in the Eastern Tropical North Pacific do not correspond to those at the California margin, suggesting that California margin oxygen signals were influenced by intermediate water from the north, namely, NPIW [Hendy and Pedersen, 2006].

[4] NPIW forms today in the Oyashio and Kuroshio region east of Japan [Talley et al., 1995], but the formation


All supporting information may be found in the online version of this article.

Received 25 June 2012; revised 29 October 2012; accepted 31 October 2012; published 23 March 2013.
and circulation of this water mass in the past is poorly understood. High $\delta^{13}C$ at intermediate depths during the last glacial suggests that the Bering Sea may have been an area of more active intermediate water formation in the past [Gorbarenko, 1996; Keigwin, 1998; Matsumoto et al., 2002], potentially helping to link high latitude climate change to oceanographic changes throughout the North Pacific. To investigate the role of NPIW circulation in millennial-scale climate change in the North Pacific, we generated multiproxy, high-resolution records of past conditions at the Integrated Ocean Drilling Program (IODP) site U1340 in the Bering Sea.

2. Study Area

[5] The Bering Sea has high productivity related to enhanced vertical mixing at the shelf break and iron-limited conditions in the open sea [Walsh et al., 1989; Banse and English, 1999; Aguilar-Islasa et al., 2007] where Site U1340 is located. The primary source of surface water entering the Bering Sea is the relatively warm Alaskan Stream, which enters the basin between the Aleutian Islands and contributes to the gyre circulation of the basin (Figure 1). Deep and intermediate water transport in/out of the Bering Sea basin occurs through the deeper straits (Figure 1) [Stabeno et al., 1999].

[6] Temperature, salinity, and oxygen profiles from the Bering Sea show that NPIW lacks a strong expression in the basin today, and the depth of the 26.8 $\sigma_t$ surface (at which NPIW is typically centered) is quite shallow (~300 m depth) [Roden, 1995]. However, chlorofluorocarbons content indicates that a small amount of ventilation has occurred in the recent past [Warner and Roden, 1995]. There is also paleoceanographic evidence that Bering Sea intermediate water distribution may have been different in the past; for example, radiolarian assemblage data indicate the presence of cold, well-oxygenated intermediate water at deeper water depths during the last glacial period and some changes over millennial timescales [Wang and Chen, 2005; Tanaka and Takahashi, 2005; Itaki et al., 2009].

3. Materials

[7] This study uses sediment cores collected during IODP Expedition 323 [Takahashi et al., 2011] on Bowers Ridge (Figure 1). The water depth at U1340 is 1324 m in the present-day oxygen minimum zone (OMZ) (Figure 1). Samples taken from the top 24 m of U1340A and U1340D were used to construct spliced continuous records; the splice (see auxiliary material) was constructed based on the correlation of shipboard magnetic susceptibility data.

[8] U1340 sediment cores contain alternating layers of diatomaceous ooze and pelagic mud and include eight distinct laminated intervals (see auxiliary material). Diatomaceous ooze, both laminated and nonlaminated, is composed of low-density, well-preserved diatom valves with grain sizes >15 μm, whereas pelagic mud is denser, finer grained (<15 μm), and mainly composed of less well-preserved diatom valves with some fine siliciclastics [Takahashi et al., 2011; Aiello and Ravelo, 2013]. More than 16 low-density, opal-rich intervals are present and show characteristics similar to the eight laminated sediments (Figure 2).

4. Methods

4.1. Stable Isotope Analysis

[9] Samples were taken every 5 cm down core, with additional samples taken every 1 cm in selected laminated intervals. Benthic foraminifera Uvigerina peregrina was picked from the >250 μm fraction, and planktonic foraminifera Neogloboquadrina pachyderma (left coiling) was picked from the >150 μm fraction. Foraminifera shells were analyzed for $\delta^{18}O$ and $\delta^{13}C$ isotopes using an automated common acid bath carbonate device interfaced to a Fisons Prism III dual-inlet mass spectrometer at the University of California Santa Cruz (UCSC). External precision based on replicates of Carrera Marble is ±0.08‰ for $\delta^{18}O$ and ±0.05‰ for $\delta^{13}C$. Data are expressed relative to the Vienna Pee Dee Belemnite.

[10] Bulk $\delta^{15}N$ was measured on unacidified samples taken every ~25 cm down core and every 1–5 cm through the deglaciation, early Holocene, and select laminated intervals. Bulk,

Figure 1. (Left) Bering Sea bathymetry, surface circulation (red arrows), subsurface circulation (yellow arrows), and the location of Site U1340. (Right) Vertical oxygen profile for the Bering Sea with U1340 located in the present-day OMZ. From Expedition 323 Scientists [2010].
carbonate-free, $\delta^{13}$C and C:N ratios were measured on a subset of samples where marked shifts in the bulk $\delta^{15}$N record occurred; for these analyses, sediment was acidified first, using a buffered solution of acetic acid (pH = 4.7) followed by rinsing and centrifuging. Analyses used a Carlo Erba 1108 elemental analyzer interfaced to a Thermo Finningan Delta Plus XP IRMS at UCSC. External precision based on replicate analyses is $\pm 0.20\%$ for $\delta^{15}$N, $\pm 0.12\%$ for $\delta^{13}$C, and $\pm 0.36$ for C:N ratios.

4.2. Percentage of Clay Mineral Analyses

Smear slide analyses of the composition of bulk sediment provided an estimate of the percent of clay minerals. Each slide was randomly counted 10 times using a Leica light transmitting petrographic microscope. The bulk sediment sample was categorized into 11 main components specific to the Bering Sea hemipelagic sediments of this study, and the percent abundance for each was estimated (Rothwell 1989), averaged over the 10 counts, and normalized based on the average total percent cover of that sample. The average standard error is 0.34%. Only the clay mineral $\%$ is used in this study to assess, along with the $\delta^{13}$Corg and C:N values, the possible influence of clay-bound inorganic nitrogen on the bulk $\delta^{15}$N record.

4.3. Alkenone Analysis

SSTs were estimated for select laminated intervals, the deglaciation and the early Holocene using alkenone paleothermometry. Lipids were extracted from 2 to 3 g of sediment with dichloromethane using a Dionex ASE 200.
dried with a nitrogen stream, and then redissolved in 150 μL of toluene spiked with hexatriacontane and heptatriacon- tane internal standards. Alkenones were quantified using a Hewlett-Packard 6890 GC-FID at UCSC following the protocol used by Dekens et al. [2007]. The reproducibility of liquid consistency standards was ±0.007 U137 units, and that of a Bering Sea sediment standard from IODP site U1342 was ±0.028 U137 units. SSTs were calculated from U137 using the Müller et al. (1998) calibration.

5. Age Model

[13] For radiocarbon analyses, benthic foraminifera U. peregrina (and in some cases other species) (Table 1) and mixed planktonic foraminifera were picked from the >250 and >150 μm fractions, respectively. Initially, samples were sonicated to remove fine particulates, but this practice caused significant destruction of planktonic tests and was discontinued. Next, the samples were lightly leached in weak HCl, rinsed, and dried on a heating block. Radiocarbon dating

![Figure 3](image-url). Comparison of U1340 benthic δ18O to the LR04 [Lisiecki and Raymo, 2005] global benthic stack. In addition to the radiocarbon dates in the upper section of core, the U1340 age model incorporates a tie-point to the LR04 stack at 62 ka.

![Figure 4](image-url). U1340 apparent ventilation age compared with core photos, planktonic δ18O, bulk δ15N, and sediment density. Modern-day ventilation age indicated by a yellow diamond. Yellow bars mark laminated intervals. Note that changes in the apparent ventilation ages could be a result of changes in the local ΔR.
was performed at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory. Ages are expressed in conventional 14C years BP using the Libby half-life of 5568, and results include a background subtraction based on 14C-free calcite and measured δ13C (cf. Stuiver and Polach, 1977). The possible contamination of foraminifera shells by authigenic mineral coating was considered to be insignificant based on visual microscopic inspection of very clean, pristine shells and on δ13C and δ18O analyses performed on splits (Table 1), which do not indicate anomalous values.

For our age model (Table 1), radiocarbon ages were converted into calendar years before present with Calib 6.0.1 [Stuiver et al., 2005], using the Marine09 calibration data set [Reimer et al., 2009] and incorporating a global ocean reservoir correction (R) of 400 years. A constant ΔR of 375 years [Robinson and Thompson, 1981; Dumond and Griffin, 2002; Kuzmin et al., 2001; Koven and Easterbrook, 2002; Cook et al., 2005; Kuzmin et al., 2007; Caisse et al., 2010] was applied before calibration. We incorporate a large uncertainty in ΔR (±300 years) because reservoir age in this region is poorly constrained and likely fluctuated when deep and intermediate water circulation underwent significant changes [Okazaki et al., 2010; Lund et al., 2011; Thornalley et al., 2011]. Dates between 14C ages were linearly interpolated. In addition, we incorporate one tie point to the Lisiecki and Raymo [2005] benthic δ18O stack at the beginning of marine isotope stage (MIS) 3 (Figure 3).

Sedimentation at U1340A averages ~30 cm/kyr, with low values during MIS 3 (22 cm/kyr), higher values at the Last Glacial Maximum (LGM) (51 cm/kyr), initial deglaciation (60 cm/kyr) and laminated BA (32 cm/kyr) and pre-Boreal (25 cm/kyr), and then low values during the YD (12 cm/kyr) and Holocene (15 cm/kyr).

6. Results
6.1. Ventilation Age

Apparent ventilation (Δb-p) was reconstructed from the radiocarbon age difference (in 14C years) between co-occurring benthic and planktonic samples (Table 1, Figure 4). Modern Δb-p in the eastern Bering Sea is 1200 years (estimated using a Δ14C profile [Key et al., 2004] to calculate 14C at depth and R + ΔR to estimate surface 14C). Δb-p was high during the early deglaciation (~9100 years) but fell to ~1000 years just before the BA. Two measurements during the late BA and the YD show an even further reduction to ~350 years, but by the end of the pre-Boreal laminated interval, Δb-p had risen to ~2400 years. However, if a change in ΔR (see section 7.1.1) explains the apparent early onset of the BA, then the relatively low Δb-p values during the deglaciation

Figure 5. All U1340 isotope records generated for this study, with gamma ray attenuation (GRA) bulk density [Expedition 323 Scientists, 2010]. Yellow bars mark laminated sediments.
(Table 1, Figure 4) are an estimate of the minimum ventilation age (see section 7.1.4).

6.2. Oxygen Isotopes

[17] The benthic δ18O record (Figure 5) has a pronounced drop of −0.75‰ at the beginning of MIS 3 (~60ka) that lasted until about 35 ka before returning to the pre-Holocene average of ~4.5‰. This is followed by a rapid increase to ~5‰ at ~16 ka, with high values persisting until the onset of BA lamination, when δ18O fell sharply. A final drop to the Holocene average of ~3.75‰ occurred at the beginning of the pre-Boreal lamination.

[18] The reduction in the benthic δ18O record at MIS 3 is also obvious in the planktonic record (Figure 5). Low values persisted until the deglaciation, when δ18O rose sharply by ~1.1‰ to ~4‰, and then fell abruptly to 3.1‰ and continued to decline more gradually through the BA and pre-Boreal laminated periods. Millennial-scale low δ18O peaks occurred during MIS 3–4, mostly corresponding to laminated opal-rich intervals. The lack of low δ18O peaks during nonlaminated opal-rich intervals may be due to lower sampling resolution and bioturbation.

6.3. Carbon Isotopes

[19] Benthic δ13C values were lowest (~1.5‰) in the oldest portion of the record, but increased to higher values at the beginning of MIS 3, coeval with the reduction in δ18O (Figure 5). Values averaged ~1.1‰, until the deglaciation when δ13C dropped back to ~1.5‰. During the BA, δ13C values increased slightly, peaking at ~0.75‰, then dropped to ~1.2‰ during the YD and pre-Boreal laminated periods. Holocene values are around ~0.75‰.

[20] Planktonic δ13C trends largely replicate benthic δ13C (Figure 5) with a marked increase in values at the beginning of MIS 3. A maximum of ~0.2‰ occurred at 51 ka and remained elevated until the deglaciation. A brief negative excursion to ~0.2‰ occurred at ~16 ka and persisted for 800 years, after which δ13C rose again to ~0.3‰. During the BA and YD, δ13C rose slowly reaching average Holocene values of ~0.5‰. Millennial-scale low δ13C spikes of relatively small amplitude (~0.2–0.5‰) occurred during MIS 3–4 in most of the laminated, opal-rich intervals. The lack of light δ13C values during nonlaminated opal-rich intervals may be due to lower sampling resolution and bioturbation.

6.4. U1340 Bulk δ15N

[21] Bulk δ15N was relatively high in the oldest part of the record, dropped at the beginning of MIS 3, and began rising again at ~27 ka (Figure 5) to values ~5.5‰ during the LGM and early deglaciation. During the BA, δ15N rose rapidly, peaking at 8.24‰, decreased to values of ~6.7‰ through the late BA and the YD, and fell to and stabilized at the Holocene average of ~4.4‰ midway through the pre-Boreal laminations. Sharp δ15N peaks occurred during three laminated intervals, with magnitudes of 1.5–2‰ and maximum values of 6–7‰ (Figure 5). The lack of dramatic δ15N peaks during nonlaminated low-density intervals may be due to lower sampling resolution and bioturbation. To assess whether changes in contributions from terrigenous sources of organic and inorganic nitrogen affected bulk δ15N values, they were compared with δ15Corg and C:N values and to smear slide estimates of the percentage of clay. Linear regression analyses indicate weak, but significant (p<0.05) correlation of δ15N with δ13Corg (R2=0.17) and with clay % (R2=0.11), but not with C:N values.

6.5. U1340 Alkenone Temperatures

[22] The concentrations of alkenones were generally low, probably due to low abundances of alkenone-producing Emiliania huxleyi in the Bering Sea [Takahashi et al., 2000]. However, it has been shown that U1340-based temperatures are consistent with observed spring/fall surface and mixed layer temperatures [Shin et al., 2002; Harada et al., 2003], and we were able to quantify U1340 in some intervals. SST increased from ~4°C during stadials (nonlaminated intervals) to ~9°C during interstadial (laminated) events. Temperatures averaged ~6°C during the LGM, and deglacial warming began around 19 ka (Figure 6) initially reaching ~8.5°C. BA warming began with the onset of laminated sediments and peaked at 11.2°C. Temperatures fell briefly to ~8°C during the YD then rose to 10.7°C during the pre-Boreal lamination. Holocene temperatures averaged ~7.7°C.

7. Discussion

7.1. Bering Sea Paleoenvironment Since the Last Glacial Maximum

7.1.1. Chronological Issues

[23] Deglacial sediments at U1340 feature two low-density laminated sections separated by a brief bioturbated
interval (Figure 4), similar to many well-dated Bering Sea and North Pacific cores with laminated intervals dating to the BA and pre-Boreal warm periods on either side of bioturbated sediment of YD age [Behl and Kennett, 1996, Zheng et al., 2000; Cook et al., 2005; Ikehara et al., 2006; Brunelle et al., 2007; Ishizaki et al., 2009; Kim et al., 2011]. This provides convincing evidence that we have correctly identified the BA (see section 7.1.3). However, our 14C age model (which assumes a constant ΔR) indicates that the onset of laminated conditions occurred at ~13.3 ka, 600 years before the start of the BA in well-dated Greenland ice cores [Alley et al., 1993] and in California [Hendy et al., 2002] and Alaskan [Davies et al., 2011] margin cores.

[24] Evidence for similar apparent early warming has been observed in other records [Seki et al., 2002; Kiefer and Kienast, 2005; Sarthein et al., 2006], where calibrated radiocarbon ages (assuming no changes in ΔR) were used and the early warming was assumed to be real. We suggest that a more reasonable explanation for the apparent lead in these and our records is local changes in ΔR. Several northeastern Pacific studies have calculated unusually old reservoir ages (1100–1800 years) during the deglaciation, BA, and YD [Kovar and Easterbrook, 2002; Cosma et al., 2008], in contrast to modern reservoir ages of ~800 years. By assuming that the laminated BA interval at U1340 occurred concurrently with the BA in Greenland ice cores and with a well-dated Alaskan margin core [Davies et al., 2011], the local reservoir age during this period was estimated to be 1300–1400 years (within the range estimated by Kovar and Easterbrook [2002] and Cosma et al. [2008]). The apparent early onset of the BA at U1340 and other subarctic Pacific sites may thus simply be the result of uncertainties and underestimation of reservoir age.

7.1.2. The LGM and the Initial Deglaciation

[25] The initial period of deglaciation at U1340 is characterized by a warm-cold pattern very similar to that observed in other subarctic Pacific records (Figure 6). Rapid warming occurred at 19.3 ka, followed by cooling from 16 to 15.3 ka, just before the BA. This pattern was noted by Kiefer and Kienast [2005] in other subarctic Pacific records of the deglaciation; small differences in timing between records can be explained by uncertainties in the radiocarbon age models and reservoir age assumptions for each of these studies.

[26] During the LGM and initial period of deglaciation, planktonic δ18O values were >1.5‰ higher than in the Holocene (Figure 6); this can be attributed to a combination of global ice volume changes [e.g., Fairbanks, 1989] and relatively high local δ18O and/or cool temperatures consistent with a southward expansion of sea ice in the LGM [Tanaka and Takahashi, 2005] and lasting until 16 ka. Benthic δ18O trends after 16 ka are more akin to a (mostly) ice volume component and track well with the Lisiecki and Raymo [2005] benthic stack (Figure 3).

7.1.3. BA, YD, and Early Holocene

[27] Throughout the North Pacific, the BA warm period (14.7–12.8 ka) was characterized by reduced subsurface oxygenation and the preservation of laminations [Kennett and Ingram, 1995; Behl and Kennett, 1996; Cannariato and Kennett, 1999; Zheng et al., 2000; Cook et al., 2005; Ikehara et al., 2006; Brunelle et al., 2007; Ishizaki et al., 2009, Kim et al., 2011]. Changes in oxygenation arose from changes in primary productivity [Mix et al., 1999; Stott et al., 2000; Crusius et al., 2004; Dean, 2007], NPIW ventilation [Duplessy et al., 1988; Kennett and Ingram, 1995; Zheng et al., 2000; Ahagon et al., 2003; Sagawa and Ikehara, 2008], or some combination of the two [Hendy and Pedersen, 2005; Ishizaki et al., 2009, Kim et al., 2011].

[28] At U1340, relatively high productivity during the BA is inferred from low sediment density (Figure 6), indicative of higher biogenic opal content [Takahashi et al., 2011] from the ubiquitous well-preserved whole diatom frustules compared with the highly fragmented diatom valves dominating the massive sediment intervals before and after the BA [Aiello and Ravelo, 2013] and from the occurrence of Chaetoceros resting and vegetative valves (Table 2). There is also a sharp increase in the abundance of Neodenticula seminae, a species indicative of Alaskan Stream waters, at the onset of lamination (Table 2) likely due to rising sea level and enhanced flow into the Bering Sea, which could have contributed to the stronger Bering Slope Current. Kim et al. [2011] suggest that a stronger Bering Slope Current and increased glacial meltwater led to a greater supply of nutrients and thus high productivity during the BA. Increased (iron) availability, perhaps due to rising sea level, is thought to have caused enhanced productivity in the Gulf of Alaska [Addison et al., 2012] and may also be the case in the Bering Sea. Our results support the idea of Crusius et al. [2004] that high productivity in the western subarctic Pacific during the BA could explain the observed North Pacific–wide subsurface oxygen minima without invoking reduced ventilation.

Table 2. Key Diatom Species Percentages Through the Transition From Bioturbated Deglacial to Laminated BA Sediments

<table>
<thead>
<tr>
<th>Sediment type</th>
<th>Depth (mcd)</th>
<th>Chaetoceros sp. a</th>
<th>Actinocyclus curvatulus (associated with postbloom conditions)</th>
<th>Neodenticula seminae (Alaskan stream indicator)</th>
<th>Thalassiosira antarctica RS (sea ice-related species)</th>
<th>Thalassiosira trifilata group</th>
<th>Fragmented valves, % total</th>
<th>Fragmented valves counted</th>
</tr>
</thead>
<tbody>
<tr>
<td>Laminated</td>
<td>1.5445</td>
<td>17.41</td>
<td>0</td>
<td>55.72</td>
<td>0</td>
<td>16.42</td>
<td>16.92</td>
<td>201</td>
</tr>
<tr>
<td></td>
<td>1.6045</td>
<td>36.22</td>
<td>1.02</td>
<td>25.51</td>
<td>2.55</td>
<td>24.49</td>
<td>28.57</td>
<td>196</td>
</tr>
<tr>
<td></td>
<td>1.6545</td>
<td>3.49</td>
<td>1.31</td>
<td>77.33</td>
<td>0</td>
<td>12.66</td>
<td>13.97</td>
<td>229</td>
</tr>
<tr>
<td></td>
<td>1.7045</td>
<td>56.28</td>
<td>0</td>
<td>21.31</td>
<td>0.5</td>
<td>4.52</td>
<td>6.03</td>
<td>199</td>
</tr>
<tr>
<td></td>
<td>1.8045</td>
<td>25.37</td>
<td>1.49</td>
<td>38.81</td>
<td>0</td>
<td>5.47</td>
<td>12.44</td>
<td>201</td>
</tr>
<tr>
<td>Transitional</td>
<td>1.8445</td>
<td>58.21</td>
<td>1</td>
<td>9.95</td>
<td>0.5</td>
<td>11.44</td>
<td>12.94</td>
<td>201</td>
</tr>
<tr>
<td>Bioturbated</td>
<td>1.9495</td>
<td>8.94</td>
<td>11.38</td>
<td>9.76</td>
<td>17.89</td>
<td>26.02</td>
<td>59.35</td>
<td>123</td>
</tr>
<tr>
<td></td>
<td>2.1495</td>
<td>1.45</td>
<td>30.43</td>
<td>11.59</td>
<td>27.54</td>
<td>20.29</td>
<td>81.16</td>
<td>69</td>
</tr>
</tbody>
</table>

aContains both vegetative and resting spores.
[30] The onset of the BA was also characterized by a rapid 5°C warming that peaked at ~11°C (Figure 6) and a sharp reduction in the resting spores of sea ice-related diatom species *Thalassiostra antarctica*, implying reduced spring sea ice cover (Table 2). Planktonic δ¹⁸O decreased abruptly ~130 years before the rise in temperature, most likely indicating freshening along with the effect of reduced global ice volume.

[31] Bulk δ¹⁵N at U1340 rose sharply to 8.24‰ during the early BA (Figure 6). One possibility is that terrigenous organic and clay-bound inorganic nitrogen influence the bulk δ¹⁵N composition in U1340, as they do in Arctic sediments [Schubert and Calvert, 2001]. In fact, at U1340, the BA and other opal-rich laminated intervals, have lower clay mineral %, higher δ¹⁳Corg, and higher C:N values compared with the massive sediment below and above such intervals. However, the lack of a statistically significant correlation between bulk δ¹⁵N and C:N values and the weak correlations between bulk δ¹⁵N and δ¹³Corg and between bulk δ¹⁵N and clay mineral % (see section 6.4) suggest that although the bulk δ¹⁵N record may be secondarily influenced by terrigenous sources, it primarily reflects changes in the δ¹⁵N of marine-derived organic matter. This is expected on Bowers Ridge due to the high productivity of the Bering Sea, unlike the less productive Arctic Ocean [Schubert and Calvert, 2001].

[32] The sharp rise in bulk δ¹⁵N at U1340 during the early BA (Figure 6) can be explained by the intensification of local/regional denitrification in the Bering Sea. Although productivity likewise increased, higher nitrate utilization cannot fully explain this rise to values of 8.24‰. When surface nitrate is completely used, the accumulating particulate organic matter bears the isotopic signature of the source nitrate [Sigman et al., 2009], which is ~5–6‰ in the Bering Sea today [Lehman et al., 2005]. It is thus unlikely that enhanced nitrate utilization produced the 7–8‰ values of the BA without a concurrent change in the δ¹⁵N of the nitrate source. In addition, changes in utilization typically affect both δ¹⁵N and δ¹³C of dissolved inorganic carbon (DIC), and U1340 values of δ¹³C of planktonic foraminifera (which reflect the δ¹³C of DIC) were constant through the BA (Figure 5). This suggests that the U1340 δ¹⁵N signature was caused by water column denitrification, which significantly raises δ¹⁵N [Sigman et al., 2009]. Our data are consistent with another Bering Sea study using core JPC-17 that measured 8–9‰ peaks in bulk δ¹⁵N during the BA [Brunelle et al., 2007], although the preceding glacial δ¹⁵N values are much lower in U1340 than in JPC-17 (see section 7.2). Differences in bulk δ¹⁵N records from sites so closely located could be explained by differences in the relative influence of marine and terrigenous sources of nitrogen, but there is ample evidence from sites around the North Pacific that there was a widespread rise in denitrification during the BA [Emmer and Thunell, 2000; Kienast et al., 2002; Kao et al., 2008; Brunelle et al., 2010; Addison et al., 2012]. The denitrification signature at U1340 may have been transmitted to the Bering Sea via intermediate water originating elsewhere in the region, but the laminations at U1340 suggest that oxygenation was low enough for water column denitrification to occur locally.

[33] The coherent low-density BA interval is composed of two distinct laminated sections separated by 23 cm (~60 years) of organic-rich bioturbated sediments (Figure 6). This nonlaminated layer may represent the brief Older Dryas cool period, which occurred between the BA events, or the Inter-Allerød Cold Period, an even shorter cool interval during the Allerød [Benson et al., 1997]. This lapse in laminations may be indicative of a slight reduction in productivity, such that oxygen was not completely depleted at depth.

[34] The YD is not a prominent feature in the U1340 records and is identified as an 11 cm thick massive interval following the laminated BA interval and preceding a second, pre-Boreal laminated interval (Figure 4). Sedimentation rate was ~10 cm/kyr during the YD interval, which is thus represented by only a few data points and likely includes bioturbated BA-aged organic material. Relatively high planktonic δ¹⁸O values and one relatively cool alkenone SST data point suggest a cooling of 1–2°C during the YD (Figure 6).

7.1.4. Intermediate Water Ventilation During the Deglaciation and Early Holocene

[35] In the North Atlantic, an inferred reduction in North Atlantic Deep Water formation occurred during cold periods (e.g., Heinrich Event 1) [Keigwin and Lehman, 1994; McManus et al., 2004], whereas ventilation increased during the BA and the Holocene [Robinson et al., 2005; Thrornalley et al., 2011]. In contrast, North Pacific data suggest that intermediate-depth ventilation was stronger during cold periods [Duplessy et al., 1988; Ahagon et al., 2003; Sagawa and Ikeharas, 2008], consistent with a modeling study by Mikolajewicz et al., 1997]. Okazaki et al. [2010] suggest that this antiphase pattern of overturning circulation could have been important in maintaining poleward heat transport when NADW formation was reduced. Taken at face value, our Δb-p apparent ventilation reconstruction differs significantly from the antiphase model proposed by Okazaki et al. [2010], however, if the ΔR changed, as discussed in section 7.1.1, then our Δb-p values may not be an accurate reflection of changes in ventilation. Thus, we offer two possible approaches to interpreting the Δb-p values.

[36] The first possibility is that the ΔR did not change significantly, and thus the low Δb-p values provide evidence for lower-than-modern ventilation during the BA and YD. U1340 apparent ventilation age was lower than modern in the BA samples, and in the YD samples, which likely incorporated, through bioturbation, a significant amount of BA-age foraminifera (see section 7.1.3.) (Figure 4). These ventilation values are typical of regions with active overturning and well-ventilated intermediate/deep waters [Matsumoto, 2007], and occur at U1340 when North Atlantic overturning was also strong. Furthermore, the highest ventilation ages at U1340 occurred during the Holocene (~2500 years) and early deglaciation (9000 years). This interpretation of the U1340 Δb-p data differs with the model of antiphased circulation [e.g., Okazaki et al., 2010]. It also strongly supports the idea that high productivity in NPIW source regions could decouple NPIW oxygenation and ventilation during the BA [Crusius et al., 2004] and that laminations during this period were caused primarily by increased productivity, rather than poor intermediate water ventilation.

[37] The second possibility is that the ΔR was ~600 years higher during the BA compared with before (the glacial period) and afterward (the Late Holocene and modern ocean) (see section 7.1.1) due to the upwelling of relatively radiocarbon depleted subsurface water. In this case, the reduced Δb-p values during the BA may be related to the increase in ΔR, although deepwater radiocarbon content was nearly constant through the deglaciation. Future work to reconstruct
the spatial patterns of ΔR and apparent ventilation age changes throughout the Bering Sea and North Pacific will provide the context to determine which oceanographic processes might explain the Δb-p values measured at U1340 during the deglaciation. [37] The Δb-p value of 9000 years at 17.8 ka (Figure 4) is reminiscent of the anomalously old intermediate and abyssal ventilation data from elsewhere in the Pacific [Sikes et al., 2000; Marchitto et al., 2007; Stott et al., 2009]. These anomalies may have resulted from the release of an isolated 14C-depleted carbon reservoir from the deep ocean during the “Mystery Interval” [17.5–14.5 ka, Denton et al., 2006] when atmospheric Δ14C dropped by ~19‰ [Beck et al., 2001; Hughen et al., 2004; Fairbanks et al., 2005; Broecker and Barker, 2007]. However, there is no evidence for an anomalously old water mass in the equatorial or South Pacific [Broecker et al., 2004; Broecker et al., 2008; De Pol-Holz et al., 2010], and modeling [Hain et al., 2011] suggests that an abyssal, radiocarbon-deplete reservoir is unlikely and insufficient to explain the intermediate-depth radiocarbon anomalies where they have been observed. Although the existence of this old carbon reservoir thus remains an open question, U1340 data may suggest an Arctic or northern source for any such low 14C water mass.

7.2. Intermediate Water Formation During MIS 3

[38] At ~60 ka, there are pronounced shifts in all the U1340 stable isotope records toward lower values in planktonic and benthic δ18O and bulk δ15N and higher values in planktonic and benthic δ13C (Figure 5). These trends are absent from California margin and North Atlantic δ18O records (Figure 7) [McManus et al., 1999; Hendy et al., 2002; Hendy and Kennett, 2003; Lisiecki and Raymo, 2005] and from North Pacific δ15N records (Figure 8) [Emmer and Thunell, 2000; Kienast et al., 2002; Hendy et al., 2004; Brunelle et al., 2010], implying that U1340 records reflect regional or local processes rather than global trends. There may have been changes in either biogeochemical cycling (e.g., productivity, nutrient/particulate organic matter fluxes, etc.) or in preformed water mass characteristics throughout the water column at the site. The first possibility is unlikely because sediment lithology does not change, as indicated by the sediment density record (Figure 5) and a nearby biogenic opal accumulation record [Brunelle et al., 2007]. Also, the benthic and the planktonic δ13C trends are in the same direction and of the same large magnitude (0.6–0.75‰; Figure 5), whereas a change in carbon flux should produce opposing trends. Finally, it is unclear how a biogeochemical process could have affected the planktonic and benthic δ18O records to such a degree. Rather, the shifts in isotope records at ~60 ka likely reflect changes in the preformed water mass characteristics affecting site U1340.

[39] Higher δ13C values after 60 ka indicate a greater contribution of low-nutrient, high-O2 water. The decrease in δ18O indicates that this water mass was either warmer by 3–4 °C or had lower 18Owater, which would translate to lower salinity. Limited U1340 alkenone data show no significant change in baseline temperatures, and Bering Sea radiolarian assemblages (discussed below) indicate the presence of cold intermediate water during the last glacial. Thus, the shift toward lower planktonic and benthic δ18O values probably reflects a decrease in salinity in the surface and at depth.
Together, U1340 records indicate the presence of a cold, low-salinity, well-oxygenated intermediate water mass, implying that NPIW could have been more dominant and extensive from 60 ka until the deglaciation (MIS 3–2) compared with the modern ocean. This idea is supported by Rella et al. [2012], whose benthic δ18O data show that Bering Sea intermediate water became colder and/or more saline than intermediate water in the Okhotsk Sea during particularly cold periods in the last glacial. These authors interpret these data as an indication that the zone of active intermediate water formation moved from the Okhotsk to the Bering Sea during these periods.

The presence of newly formed intermediate water in the Bering Sea during MIS 3–2 is supported by the abundances of radiolarian Cycladophora davisiiana in Bering Sea sediments. Today, C. davisiiana is abundant (>20%) only in the Okhotsk Sea, particularly in cold, low-salinity intermediate water at depths of 300–1000 m, and is sustained by high microbial biomass supported by high fluxes of particulate organic carbon related to the formation of seasonal sea ice and brine [Wimmergut and Abelmann, 2002; Okazaki et al., 2003]. High relative abundances of C. davisiiana in the Bering Sea during MIS 2–3 support the presence of NPIW-like water [Tanaka and Takahashi, 2005] and provide a mechanism (increased seasonal sea ice and brine formation) for local intermediate water formation. Rella et al. [2012] further suggest that this increased sea ice extent and brine formation was driven by atmospheric changes, leading to stronger northerly winds during cold periods in the last glacial.

The U1340 δ15N record is also consistent with increased local intermediate water formation ~60 ka. Because there is not a strong correlation between the bulk δ15N values and δ13Corg, C:N, or percentage of clay minerals, we interpret the δ15N signal as primarily reflecting changes in oceanographic conditions rather than changes in the delivery of terrigenous nitrogen to the site. Thus, the rapid 2% drop in bulk δ15N (Figure 5) is most likely related to reduced nutrient utilization [Altabet and Francois, 1994] caused by a decrease in productivity or an increase in nitrate availability via vertical mixing. The lack of a clear change in biogenic opal productivity at another Bowers Ridge site [Brunelle et al., 2007] suggests that the U1340 δ15N signal reflects higher nutrients, most likely caused by a reduction in Bering Sea stratification. Reduced stratification would provide an environment ideal for intermediate water formation, in contrast to the open northwest Pacific, where conditions were more stratified and nutrient utilization was higher [Brunelle et al., 2010; Jaccard et al., 2005]. The lack of a similar drop in δ15N values in records from outside the Bering Sea [e.g., Brunelle et al., 2010; Kienast et al., 2002; Hendy et al., 2004] (Figure 8) is compelling evidence for local rather than subarctic Pacific wide changes in stratification and intermediate water formation.

Bulk sediment δ15N and diatom-bound δ15N (δ15N_{db}) records from a different Bowers Ridge site, JPC17 (2209 m), lacks the pronounced drop observed at U1340 (Figure 8) [Brunelle et al., 2007]. However, diagenetic effects [Sigman et al., 2009] could have had a greater impact on the δ15N of the bulk sediment record at JPC17, which is deeper and more oxygenated than U1340. Furthermore, although δ15N_{db} is thought to be impervious to diagenetic alteration [Sigman et al., 1999], species-specific vital effects during biomineralization are poorly quantified, and changes in δ15N_{db} may thus incorporate the effects of shifting diatom assemblages. Lower sedimentation (~12 cm/kyr at JPC17 vs. ~30 cm/kyr at U1340) may also obscure sharp changes in JPC17 records. As described earlier, a clear, pronounced δ15N peak during the BA, following relatively low glacial δ15N values, is not recorded at JPC17 but is recorded at sites around the North Pacific with shallower water depths and/or higher sedimentation rates (U1340, this study; GGC27, Brunelle et al., 2010; 1017E, Hendy et al., 2004; W8709-8TC, Kienast et al., 2002); as such, the structure of the δ15N record at U1340, including the pronounced decrease in δ15N that occurs at ~55 ka, is most likely robust, even if it is not recorded at JPC17.

The influence of well-ventilated intermediate water at U1340 seems to have ended during the deglaciation, between 20 and 17 ka. Benthic δ18O began to rise at 20 ka, whereas planktonic δ18O rose sharply at 18 ka (Figure 5). U1340 alkenones showed no clear temperature trends during this interval (Figure 6), suggesting that the δ18O signal was caused by rising salinity. Benthic δ13C dropped from 18 to 16.5 ka, whereas planktonic δ13C fell abruptly at 18 ka (Figure 5). Changes in all of these proxies toward the typical values observed before MIS 3 suggest the end of active intermediate water formation in the Bering Sea. This change may have been influenced by increasing stratification due to warming surface temperatures and the input of glacial meltwater.

7.3. Characteristics of Millennial-Scale Events

The ~15 millennial-scale events before 20 ka at U1340 seem to be high productivity events based on their low-density, high biogenic opal content and enhanced preservation of diatom valves [Takahashi et al., 2011; Aiello and Ravelo, 2013]. Five of these events (Figure 5) were accompanied by laminations similar to those occurring elsewhere in the North Pacific during DO events [e.g., Behl and Kennett, 1996], providing striking evidence for variations in intermediate water oxygenation possibly across the North Pacific. The correlation of each millennial-scale event at U1340 with each specific DO interstadials in the Greenland ice core record was not possible, likely due to uncertainty in our age model beyond the oldest radiocarbon date (~29 ka) and to bioturbation obscuring some events when anoxia did not develop. Nonetheless, because the low-density peaks at U1340 occur over millennial timescales and number roughly the same as DO events, we refer to these intervals as “DO-type” events.

The low-density DO-type events at U1340 were characterized by warmer, fresher surface waters, as evidenced by sharp drops in planktonic δ18O (Figure 5); at the two DO-type events with the highest resolution data (Figure 9), the drop was 0.7–0.9‰, representing 3–4 °C warming if attributed solely to temperature fluctuations. This is roughly consistent with the alkenone-derived temperature record, which indicates ~3–5 °C warming during laminated intervals (Figure 9). Negative δ18O peaks could also be attributed to a freshening of surface waters due to sea ice melting and/or increased runoff into the Bering Sea. Benthic δ18O remained constant, implying that temperature/salinity changes were limited to surface waters. Our results are in contrast to a study from the Emperor Seamounts, suggesting an antiphase
pattern for DO events in the North Pacific relative to the North Atlantic [Kiefer et al., 2001].

[46] Sediment lithology and δ13C records suggest increased productivity during DO-type events, with several laminations containing relatively low planktonic δ13C values (Figure 9). During one laminated interval, the percentage of fragmented diatom valves dropped from ~75% to ~3%, indicating much higher preservation and suggesting increased opal flux, consistent with a rise in productivity. Given this productivity signal, we attribute negative δ13C peaks to amplified upwelling. Higher export production is also consistent with the reduction in benthic oxygen that is evident during DO-type events.

[47] The upwelling signature (and implied nutrient-rich environment) at U1340 during DO-type events suggests that co-occurring high δ15N peaks were caused by a change in the preformed isotopic signature of subsurface nitrate rather than increased nutrient utilization. Peak values of 5.6–6.9‰ (Figure 9) cannot be accounted for by complete nutrient utilization if subsurface nitrate retained its modern signature (5–6‰ [Lehman et al., 2005]). Further, it seems doubtful that nitrate would be completely used, given that the Bering Sea today is an HNLC, iron-limited environment with only 50% nitrate utilization [Brunelle et al., 2007]. Finally, the accompanying low planktonic δ13C values suggest lower, not higher, nutrient utilization. The δ15N peaks at U1340 were therefore most likely due to a denitrification signature in subsurface nitrate during DO-type events originating locally, given the suboxic benthic conditions, or elsewhere in the North Pacific, and suggesting possible changes in subsurface circulation during DO-type events.

8. Summary

1. The BA at U1340 was characterized by rapid warming, high primary productivity, suboxic benthic conditions, and local or regional denitrification. Our radiocarbon age model places the apparent onset of the BA at U1340 several hundred years earlier than the North Atlantic, most likely a result of large fluctuations in reservoir age in the North Pacific during the deglaciation.

2. From 60 ka through the deglaciation, there was a strong presence of cold, low-salinity, well-oxygenated intermediate water at U1340. The Bering Sea also seems to have been less well stratified than elsewhere in the subarctic Pacific during this period, suggesting that conditions were ideal for this intermediate water mass to have formed locally.

3. DO interstadials at U1340 were characterized by rapid warming and freshening, along with high upwelling-driven productivity that contributed to benthic suboxia/anoxia and led to enhanced water column denitrification (either locally or regionally).

4. Our reconstruction of Bering Sea conditions during the last 85 kyr indicates that changes in intermediate water formation and ventilation have occurred over both glacial-interglacial and millennial timescales, implying the subarctic northwest Pacific has the potential to play an active role in global climate change.

5. The anoxia and denitrification signals present at U1340 during the BA and DO events show that denitrification occurred more broadly in the past. Although beyond the scope of this study, future work should investigate if rapid changes in denitrification are an important dynamic feedback for global climate.

References


Ahagon, N., K. Ohkushi, M. Uchida, and T. Mishima (2003), Mid-depth circulation in the northwest Pacific during the last deglaciation: Evidence
Stabeno, P. J., J. D. Schumacher, and K. Ohtani (1999), The physical oceanography of the Bering Sea, in Dynamics of the Bering Sea, edited by T. R Loughlin and K. Ohtani, pp. 1–28, University of Alaska Sea Grant, Fairbanks, AK.


